Geochronology of granitic and supracrustal rocks from the northern part of the East Greenland Caledonides: ion microprobe U–Pb zircon ages

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Granitoid rocks from different settings within the northern part of the East Greenland Caledonian fold belt have yielded U–Pb zircon dates between 2000 and 1730 Ma, confirming the Palaeoproterozoic origin of the crystalline basement. Widespread sandstone sequences occur both in the Caledonian fold belt and in the foreland to the west; all of these have been assigned to the Independence Fjord Group, and attributed a Mesoproterozoic age on geological maps. However, metarhyolitic rocks associated with the sandstones in the Caledonian fold belt have yielded an age of 1740 ± 6 Ma, significantly older than anticipated. Zircon ages for a sandstone sample in the same area suggest deposition after the end of Palaeoproterozoic orogenic events, but in part prior to emplacement of the rhyolitic rocks at 1740 Ma; sandstone from another locality may have been deposited before emplacement of the latest Proterozoic granite sheets.

Field relations suggest that some granitic veins and sheets might be Caledonian in age, but, with one possible exception, all those analysed proved to be Proterozoic. The apparent absence of Caledonian granites in the northern part of the East Greenland Caledonides, despite regional high-grade metamorphism, may be related to the lack of major occurrences of pelitic supracrustal rocks within the crystalline basement complexes.

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Keywords: Caledonides, granites, Greenland, Palaeoproterozoic, SHRIMP, supracrustal rocks, zircon

The East Greenland Caledonian fold belt (Haller 1971; Henriksen & Higgins 1976; Henriksen 1985; Higgins 1995a) stretches from 70°N to 81°N, a distance of nearly 1300 km along strike. Between 76°N and 79°N the fold belt is dominated by crystalline basement complexes (Fig. 1), mainly grey gneisses and somewhat later metagranitoid rocks of Palaeoproterozoic age (gneisses *c*. 2000–1850 Ma; some granitic rocks down to about 1750 Ma; Kalsbeek *et al*. 1993; Nutman & Kalsbeek 1994). The gneisses contain local enclaves of supracrustal rocks, and are cut by several generations of mafic dykes, most of which are strongly deformed (Hull *et al*. 1994). Eclogite pods (Gilotti 1993), formed during high-grade Caledonian metamorphism (Brueckner *et al*. 1998), occur at many localities. North of *c*. 79°N the fold belt is built up of Caledonian nappe complexes involving basement gneisses together with Proterozoic and Palaeozoic supracrustal sequences (Higgins 1995b).

In this paper we report ion microprobe U–Pb zircon data on granitoid, metavolcanic and metasedimentary rocks in order to establish a broad geochronological framework for the northernmost parts of the East Greenland Caledonides. One question we wished to assess

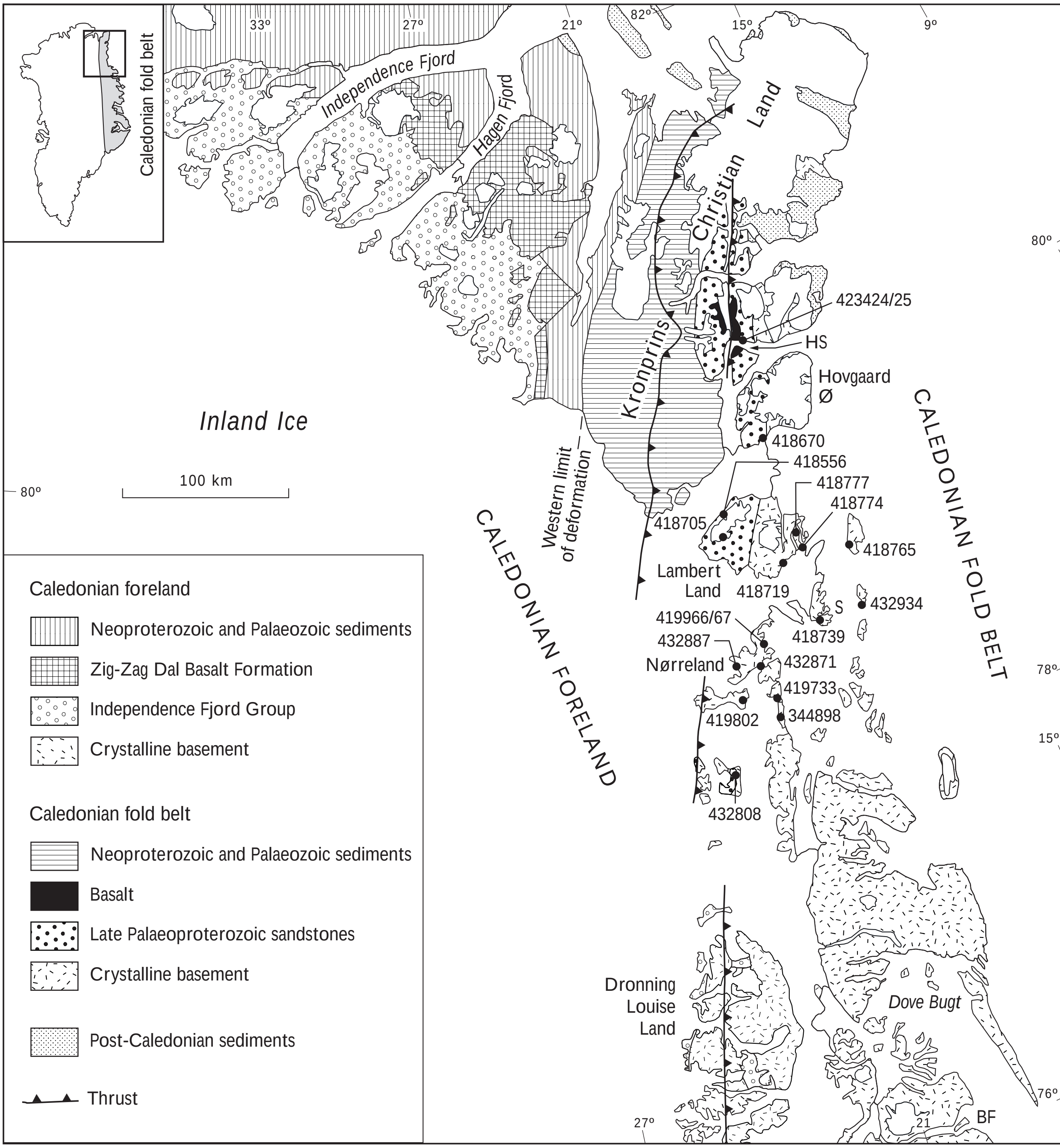


Fig. 1. Geological map of the northern part of the Caledonian fold belt and the adjoining Caledonian foreland (modified after Escher & Pulvertaft 1995). Stratigraphic correlations between the fold belt and the foreland are not entirely certain, but the late Palaeoproterozoic sandstones from the fold belt probably correlate with the Independence Fjord Group of the foreland. Correlation of the 1740 Ma basalts at Hekla Sund with the Zig-Zag Dal Basalt Formation of the foreland is unlikely. **BF**: Bessel Fjord; **HS**: Hekla Sund; **S**: Schnauder Ø. **Dots** with six digit numbers represent sample localities that refer to the files of the former Geological Survey of Greenland (GGU).

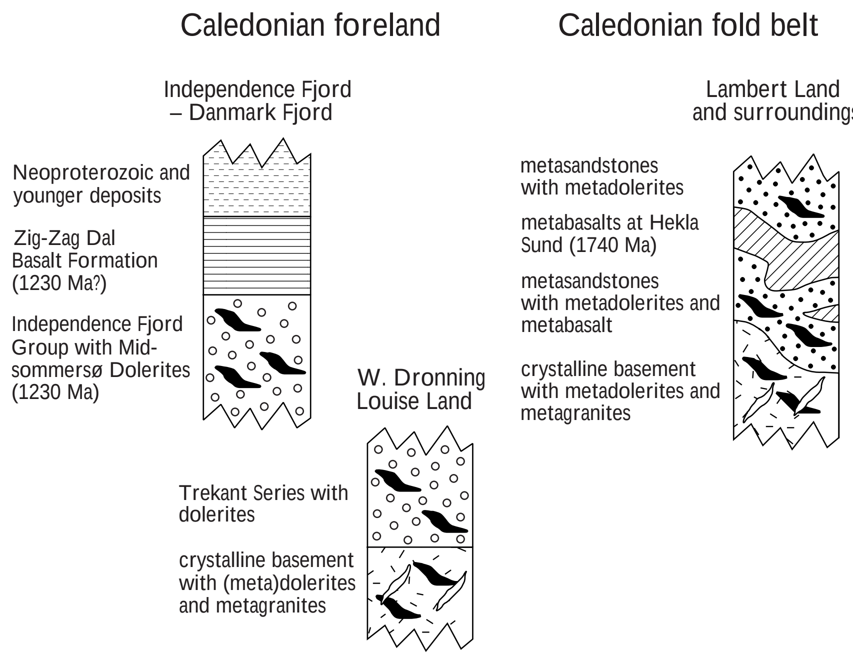


Fig. 2. Simplified stratigraphy for the Caledonian foreland and the Caledonian fold belt in North-East and eastern North Greenland.

was whether some of the granitoid rocks in the region could represent Caledonian granites; no granites of certain Caledonian age were detected by the present geochronology, however, and a possible explanation is suggested for their apparent absence in this part of the Caledonian fold belt.

Regional geology

Caledonian foreland

The foreland to the East Greenland Caledonian fold belt is widely exposed west of Kronprins Christian Land (80–81°N) and also outcrops at the margin of the Inland Ice in western Dronning Louise Land (77°N, 25°W; Fig. 1). Crystalline basement rocks are not exposed west of Kronprins Christian Land, but in North Greenland the basement consists of Archaean rocks affected by Palaeoproterozoic orogenic events, and in North-East Greenland it is made up of juvenile Palaeoproterozoic rocks (Escher & Pulvertaft 1995). Most Proterozoic orogenic activity took place 2000–1800 Ma ago; some late granites in North-East Greenland have been dated at *c*. 1750 Ma (Kalsbeek *et al.* 1993).

Between Independence Fjord and Danmark Fjord (Fig. 1) in the Caledonian foreland the lowermost unit exposed is the Independence Fjord Group, a sequence of sandstones, the base of which is not exposed (Figs 1, 2; Collinson 1980, 1983; Sønderholm & Jepsen 1991). These sandstones are undeformed and unmetamorphosed, and clearly post-date the Palaeoproterozoic orogenic events recorded in North-East Greenland. The Independence Fjord Group is cut by numerous sheets and dykes of dolerite, the Midsommersø Dolerites, for which Rb–Sr whole-rock isochron ages of *c*. 1230 Ma have been obtained (Kalsbeek & Jepsen 1983). Deposition of the Independence Fjord Group in the Caledonian foreland must thus have taken place between *c*. 1750 and 1230 Ma ago. Rb–Sr data on clay minerals from siltstone samples have suggested diagenesis at about 1380 Ma (Larsen & Graff-Petersen 1980), and on the Geological map of Greenland, 1:2 500 000 (Escher & Pulvertaft 1995) the Independence Fjord Group is consequently shown as Mesoproterozoic.

The Independence Fjord Group is overlain by the Zig-Zag Dal Basalt Formation. The age of this formation is not known, but it is believed to be contemporaneous with emplacement of the Midsommersø Dolerites (Kalsbeek & Jepsen 1984). The Zig-Zag Dal Basalt Formation is overlain by Neoproterozoic and Palaeozoic strata.

In western Dronning Louise Land the crystalline basement of the foreland to the Caledonian fold belt is unconformably overlain by a sequence of sandstones, the Trekant Series (Figs 1, 2; Peacock 1956, 1958; Strachan *et al.* 1994), which are lithologically similar to the Independence Fjord Group sandstones west of Kronprins Christian Land. Both the crystalline basement and the sandstones of the Trekant Series contain numerous dolerite sheets, and are increasingly affected by Caledonian deformation and metamorphism further east in Dronning Louise Land. No radiometric ages are available for the dolerites or the sandstones, but a broad correlation with respectively the Midsommersø Dolerites and the Independence Fjord Group is assumed (see e.g. Escher & Pulvertaft 1995).

Caledonian fold belt

The northernmost part of the East Greenland Caledonian fold belt consists mainly of gneisses and metagranitoid rocks, which are tectonically interleaved with units of quartzitic and feldspathic metasandstones traditionally correlated with the Independence Fjord Group. Sheets and dykes of metadolerite are common in all these rocks. In the area around Hekla Sund in Kronprins Christian Land (HS, Fig. 1) the metasandstones are interlayered with low-grade metamorphic pillow lavas of basaltic composition with subordinate occurrences of metarhyolitic rocks (Pedersen *et al.* 1995a, b). Pedersen *et al.* (1995a) describe syn-sedimentary faults within this sequence, and suggest that the basalts were deposited in an alluvial fan setting during subsidence along rift margin escarpments. On the Geological map of Greenland, 1:2 500 000 (Escher & Pulvertaft 1995) the basalts are correlated with the Zig-Zag Dal Basalt Formation west of Kronprins Christian Land. However, our age determinations show that this correlation is unlikely (p. 44).

A complex sequence of Caledonian thrust sheets in Lambert Land (just north of 79°N, Fig. 1) include sandstones with locally well-preserved sedimentary structures (Jones & Escher 1995). The eastern part of Lambert Land and adjoining areas consist of granitoid rocks, for some of which a Caledonian age was considered possible during field work. For example, a Caledonian age was considered likely for a complex of granites and granodiorites on Schnauder Ø (S, Fig. 1) and eastern Lambert Land; the 'Schnauder Ø complex' can be distinguished from the surrounding regional gneisses on the basis of a significantly lower degree of deformation and the lack of mafic dykes, and is shown on one of John Haller's maps as post-tectonic granite (Haller 1971, p. 267). Rafts of gneiss with remnants of mafic dykes are present within the complex. Presumed Caledonian granite sheets were also locally found within Caledonian shear zones, and interpreted as syntectonic. Granite sheets were also found cutting mafic dykes similar to the Midsommersø Dolerites, in areas mainly composed of sandstone. However, due to limited exposure and strong deformation, it was not possible to establish detailed field relationships, and nowhere were intrusive contacts observed between the granites and the sandstones. As in Kronprins Christian Land the sandstones in Lambert Land are traditionally correlated with the Independence Fjord Group of the Caledonian foreland and the metadolerites that cut the sandstones with the Midsommersø Dolerites.

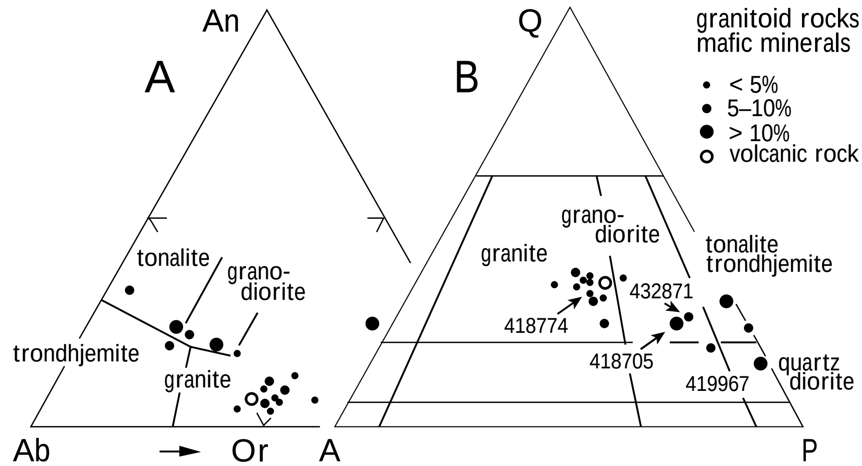
A small tectonic window in the Nørreland area (latitude 78°40′N, Fig. 1) exposes a sequence of sandstones and dolerite sheets, associated with granitoid rocks, beneath a major Caledonian thrust (Hull & Friderichsen 1995); sheared carbonates in the thrust zone have yielded Ordovician conodonts (M.P. Smith, personal communication 1998), clearly demonstrating the Caledonian age of westward thrust displacement. Field observations within the window suggest that the granitoid rocks are younger than the sandstones, although strong deformation precludes unambiguous assessment of their mutual relationships.

The northernmost part of the Caledonian fold belt was remapped by the Geological Survey of Greenland (GGU, amalgamated with its Danish counterpart in 1995 into the Geological Survey of Denmark and Greenland). The isotopic investigations reported in this paper were carried out in connection with this mapping project. Details of the field setting of the various rocks appear in an unpublished volume 'Express report: eastern North Greenland and North-East Greenland 1995' (Higgins 1995b), copies of which are available from the Survey.

Scope of the present investigation and methods used

Ion microprobe U–Pb zircon dating has been carried out on metagranitoid rocks from different settings within the fold belt: (1) orthogneisses from the crystalline basement, (2) metagranitoid sheets intruded into basement gneisses, (3) metagranitoid rocks from the tectonic window in Nørreland, (4) granitoid rocks from the Schnauder Ø complex, (5) granitoid sheets from Lambert Land and surroundings which occur within Caledonian shear zones or were intruded into mafic dykes that are lithologically similar to the Midsommersø Dolerites. In addition, zircons were studied from (6) two samples of rhyolitic rocks associated with the unit of metabasalts at Hekla Sund, and detrital zircons

Fig. 3. Classification of the investigated samples. **A**: CIPW normative An–Ab–Or diagram for granitoid rocks (O'Connor 1965, modified after Barker 1979). **B**: Q–A–P diagram (Streckeisen 1976; Q: quartz, A: alkali feldspar, P: plagioclase). Modal compositions were estimated from chemical analyses (kation-norms, with orthoclase + hypersthene recalculated into biotite + quartz: 5 or + 6 hy = 8 bi + 3 qz). The quartz dioritic sample GGU 419967 is too mafic to be represented in Fig. 3A.



were analysed from (7) two metasandstone samples, one from Lambert Land and one from the Nørreland window, to obtain an impression of the time of deposition and the source areas of the sediments.

The investigated samples vary from quartz dioritic to granitic in composition (Fig. 3). In the CIPW normative Ab–An–Or diagram (Fig. 3A; O'Connor 1965; Barker 1979) as well as in the Q–A–P (quartz–alkali feldspar–plagioclase) diagram (Fig. 3B; Streckeisen 1976; modal compositions estimated from chemical analyses) they plot as quartz diorite, tonalite, granodiorite and (leuco-) granite. The total proportion of (normative) mafic minerals is low in most cases; most granitic rocks have < 5 per cent mafic minerals.

Zircons were separated from 1–2 kg samples and mounted in 1-inch epoxy disks (20–50 grains for most samples, 5–10 samples per disk). U-Th-Pb isotopic ratios and concentrations were determined with SHRIMP 1 (SHRIMP = Sensitive High Resolution Ion MicroProbe) at the Research School of Earth Sciences, Australian National University (ANU), Canberra, using the ANU standard zircon SL13 (572 Ma; $^{206}Pb/^{238}U$ = 0.0928) for reference. Descriptions of analytical procedure and data assessment are given by Compston *et al.* (1984), Claoué-Long *et al.* (1995) and Williams (1998). As a check on the accuracy of $^{207}Pb/^{206}Pb$ ratios obtained by SHRIMP, analyses of zircons from a Palaeoproterozoic norite, QGNG, were run interspersed with unknowns. Isotope dilution thermal ionisation mass spectrometry for different QGNG zircon fractions has yielded $^{207}Pb/^{206}Pb$ ages of 1850 ± 2 Ma (C.M. Fanning, personal communication 1995) and 1850 Ma to as low as 1810 Ma (T. Skjöld, personal communication 1996). Most of the analyses of QGNG run during the present investigation yielded $^{207}Pb/^{206}Pb$ ages of 1820 to 1860 Ma, the same range as found by thermal ionisation isotope dilution analyses, with 1σ errors on individual ages typically of about 15–20 Ma.

In the context of the regional geological understanding of the study area, reconnaissance data yielding a general rather than a precisely defined age were considered sufficient for a number of samples; in these cases only three to five zircons were analysed, each spot was analysed using only four mass scans (instead of the usual five), and to compensate for the reduced counting time a larger than usual spot (50 vs. 30 µm in diameter) was employed where size and homogeneity of the zircons permitted (most granitoid rocks). Even though the precision of such age determinations is low we regard this as a useful approach as part of regional projects, since a large number of samples can be analysed in this way, providing useful results on the scale of the orogen.

Ages were calculated in two ways: (1) as the mean of the most concordant $^{207}Pb/^{206}Pb$ ages, and (2) as the upper intercept of the best-fit discordia line with concordia, using all analytical data together with an additional data point at 400 ± 50 Ma. The use of this additional data point is based on the assumption that the discordance of non-concordant data is the result of Pb loss during Caledonian metamorphism. Because most samples did not yield strongly discordant zircons, the calculations (York 1969; regressions calculated in $^{207}Pb/^{206}Pb$ vs. $^{238}U/^{206}Pb$ space assuming non-correlated errors) without this extra point yield very poorly defined lower intercepts which, however, all encompass 400 Ma as a possible time of Pb loss. A summary of results is given in Table 1 (all ages quoted with 2σ errors), and analytical data in Tables 2 and 3. For sample localities see Fig. 1 and Table 1.

Table 1. SHRIMP U–Pb zircon ages for granitoid and metavolcanic rocks from the Caledonian fold belt

GGU sample no.	Rock type	Latitude, Longitude	N spots	Age (Ma) (1)	Age (Ma) (2)	MSWD
(1) Gneisses from the crystalline basement						
344898	trondhjemitic gneiss	78°15.8′N, 21°21′W	2/3	1994 ± 38	1978 ± 112	2.2
418765	biotite gneiss	79°00.0′N, 18°00′W	2/3	1935 ± 24	1949 ± 27	0.7
418719	granitic bi. gneiss	79°02.0′N, 20°13′W	3	*c.* 1900?		
(2) Granitoid sheets intruded into basement gneisses						
419733	metagranite	78°24.8′N, 21°15′W	2/3	1751 ± 31	1769 ± 36	0.1
432934	metagranite	78°45.0′N, 18°30′W	3	1800 ± 100	1822 ± 118	0.4
419802	metagranite	78°27.8′N, 22°09′W	4			
432808	metagranite	78°01.1′N, 22°46′W	3		2013 ± 103	0.1
(3) Granitoid rocks from the tectonic window at 78°40′N						
419967	meta-quartz diorite	78°41.2′N, 21°14′W	3/5	1858 ± 35	1889 ± 42	0.6
419966	metagranite	78°41.2′N, 21°14′W	5/6	1731 ± 38	1733 ± 43	0.3
432871	metagranodiorite	78°37.4′N, 21°17′W	5	*c.* 1900	1912 ± 46	0.4
(4) Granitoid rocks from the Schnauder Ø complex						
418739	foliated granite	78°41.5′N, 19°25′W	2/3	1974 ± 18		
418777	fol. musc.-bi. granodiorite	79°08.1′N, 19°21′W	4/6	*c.* 2000		
(5) Granitoid sheets from Caledonian shear zones or intruded into metadolerites in Lambert Land						
418670	meta-augen granite	78°59.5′N, 17°58′W	6/7*	1986 ± 23	1959 ± 35	2.7
418774	musc.-bi. metagranite	79°08.8′N, 19°06′W	9#	404 ± 11	2018 ± 39	1.3
418705	meta-augen granodiorite	79°14.8′N, 21°22′W	5	1994 ± 20	1999 ± 25	0.6
(6) Rhyolitic rocks from the basalt sequence at Hekla Sund						
423424	metarhyolite	80°15.0′N, 19°13′W	9/10	1742 ± 10		
423425	metarhyolite	80°15.0′N, 19°13′W	10	1738 ± 11		

Ages were calculated in two ways: (1) as the mean of the most concordant $^{207}Pb/^{206}Pb$ ages, and (2) as the upper intercept of the best-fit discordia using all data points plus an extra point at 400 ± 50 Ma (see text).
'N spots' is the number of spots analysed and used in age calculations, for example, 3/5 indicates that 5 spots were analysed of which 3 were used for the $^{207}Pb/^{206}Pb$ age estimate and 5 for the discordia calculation.
* For 418670 two rims were not included in the discordia calculation.
\# For 418774 two spots were not included in the discordia calculation, and no extra point at 400 ± 50 Ma was employed.
Errors at 2σ.

Results

Basement gneisses

Three samples of basement orthogneisses have been studied (GGU 344898, 418765, 418719; Fig. 1). Precise ages were not considered necessary, as the main aim of the age determinations was to determine whether the rocks belonged to the same age category as samples analysed previously (1850–2000 Ma, Kalsbeek *et al.* 1993; Nutman & Kalsbeek 1994), or whether Archaean rocks were also present. Therefore only three zircons were analysed per sample. Two samples with relatively fresh (transparent) zircons (GGU 344898 and 418765) gave ages of about 2000 and 1950 Ma, respectively (Table 1). For both samples the data show some scatter along reference discordia lines with lower intercepts at 400 Ma (shown for 418765 in Fig. 4A), indicating that the discordance of non-concordant analyses can be interpreted as the result of Pb loss during Caledonian metamorphism. These results are in agreement with ages obtained earlier from the crystalline basement (Kalsbeek *et al.* 1993; Nutman & Kalsbeek 1994).

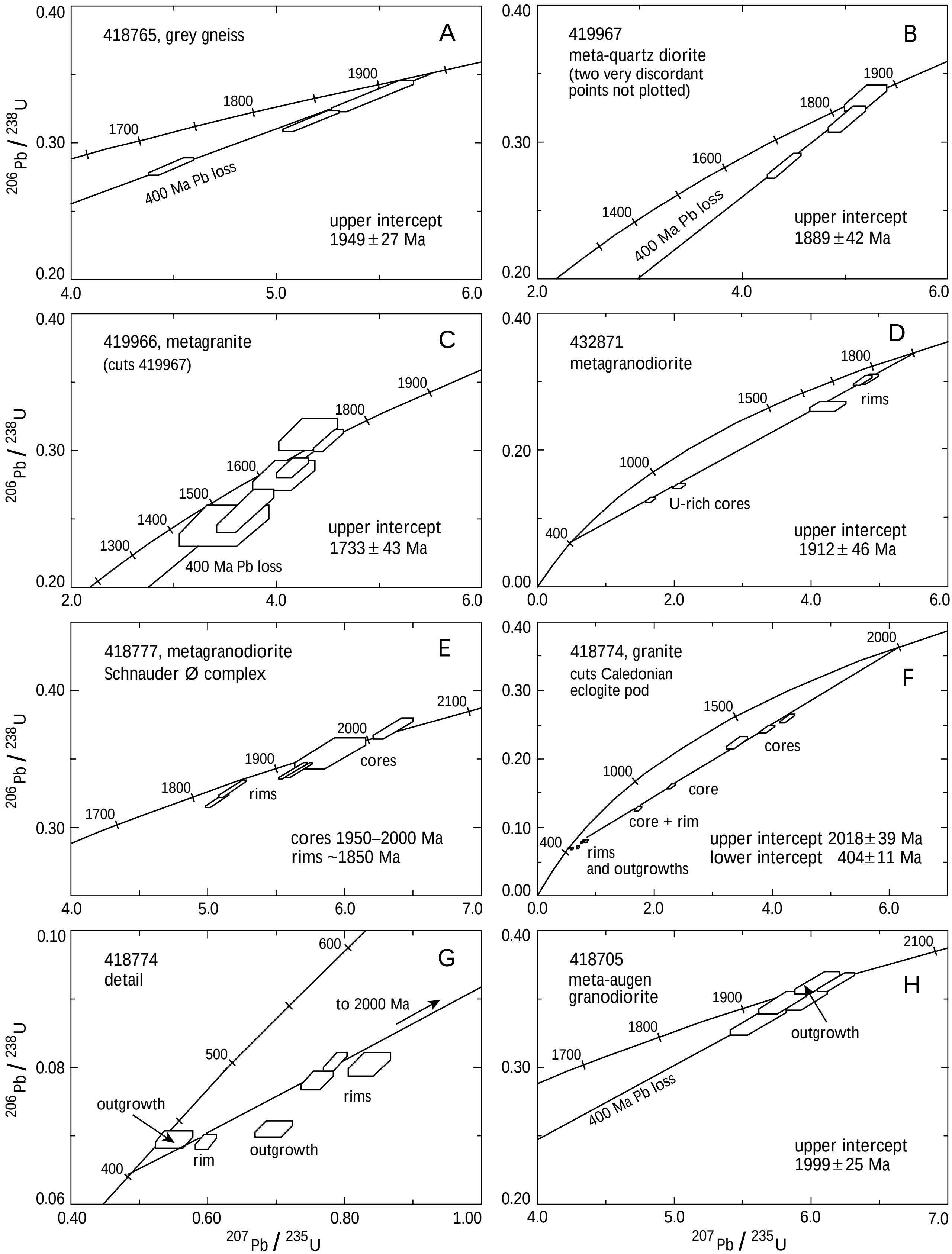


Fig. 4. Selected concordia diagrams for zircons from granitoid rocks, East Greenland Caledonian fold belt; error boxes display 1σ errors. For localities see Fig. 1. Discordia lines down to 400 Ma are for reference only.

Table 2. SHRIMP U–Pb zircon data for metagranitoid rocks from the Caledonian fold belt

Spot	U (ppm)	Th/U	f_{206} (%)	$^{206}Pb/^{238}U$	$^{207}Pb/^{206}Pb$	Age	Disc. (%)
GGU 344898							
1.1	130	0.15	0.33	0.327 ± 8	0.1208 ± 25	1967 ± 38	7
2.1	201	0.17	0.15	0.351 ± 7	0.1232 ± 15	2003 ± 22	–3
3.1	198	0.31	0.43	0.308 ± 7	0.1167 ± 14	1907 ± 22	–9
GGU 418765							
1.1	488	0.40	0.10	0.316 ± 8	0.1186 ± 10	1935 ± 14	–9
2.1	449	0.51	0.27	0.283 ± 6	0.1149 ± 10	1878 ± 16	–14
3.1	535	0.42	0.14	0.334 ± 11	0.1186 ± 13	1935 ± 19	–4
GGU 418719							
1.1	57	0.72	1.96	0.348 ± 11	0.1099 ± 48	1798 ± 81	7
2.1	245	0.78	0.32	0.268 ± 6	0.1154 ± 17	1886 ± 27	–19
3.1	330	0.11	0.69	0.216 ± 5	0.1516 ± 25	2364 ± 29	–47
GGU 419733							
1.1	412	0.40	0.19	0.285 ± 6	0.1064 ± 13	1739 ± 22	–7
2.1	328	0.28	0.24	0.297 ± 6	0.1076 ± 12	1760 ± 21	–5
3.1	452	0.23	0.42	0.196 ± 4	0.0997 ± 15	1618 ± 28	–29
GGU 432934							
1.1	65	0.93	0.48	0.310 ± 9	0.1119 ± 42	1830 ± 70	–5
2.1	44	0.76	1.26	0.265 ± 11	0.1100 ± 64	1799 ± 110	–16
3.1	77	0.90	0.67	0.278 ± 9	0.1045 ± 61	1706 ± 112	–7
GGU 419802							
1.1	175	0.50	0.56	0.342 ± 8	0.1211 ± 17	1972 ± 25	–4
2.1	649	0.57	6.86	0.119 ± 2	0.1044 ± 40	1703 ± 72	–57
3.1	138	0.30	0.89	0.284 ± 11	0.1937 ± 31	2774 ± 26	–42
4.1	1352	0.29	3.60	0.066 ± 2	0.0559 ± 30	447 ± 125	–8
GGU 432808							
1.1	2444	0.05	0.56	0.128 ± 2	0.0920 ± 8	1468 ± 17	–47
2.1	2187	0.07	0.52	0.155 ± 3	0.1007 ± 12	1636 ± 22	–43
3.1	2312	0.05	2.55	0.093 ± 2	0.0738 ± 16	1037 ± 44	–45
GGU 419967							
1.1	395	0.30	0.44	0.283 ± 9	0.1130 ± 15	1847 ± 24	–13
2.1	168	0.39	0.35	0.332 ± 9	0.1135 ± 29	1856 ± 46	0
3.1	250	0.29	0.23	0.317 ± 9	0.1150 ± 21	1880 ± 34	–6
4.1	548	0.49	1.69	0.129 ± 3	0.0980 ± 23	1586 ± 45	–51
5.1	522	0.37	1.19	0.154 ± 3	0.0995 ± 28	1614 ± 54	–43
GGU 419966							
1.1	153	1.29	1.38	0.312 ± 12	0.1004 ± 49	1631 ± 94	7
2.1	25	0.92	2.15	0.245 ± 15	0.1032 ± 106	1683 ± 204	–16
3.1	178	0.80	0.35	0.287 ± 7	0.1049 ± 26	1712 ± 46	–5
4.1	84	1.31	0.19	0.256 ± 16	0.1047 ± 38	1709 ± 68	–14
5.1	106	1.75	0.00	0.307 ± 8	0.1065 ± 13	1740 ± 23	–1
6.1	93	1.27	0.54	0.282 ± 11	0.1049 ± 61	1713 ± 110	–7
GGU 432871							
1.1*	167	0.60	0.47	0.303 ± 7	0.1155 ± 20	1888 ± 32	–10
1.2	546	0.46	1.42	0.128 ± 4	0.0940 ± 26	1508 ± 53	–49
1.3*	180	0.69	0.33	0.302 ± 8	0.1143 ± 16	1869 ± 26	–9
2.1*	235	0.33	0.16	0.264 ± 7	0.1164 ± 60	1902 ± 95	–21
3.1	551	0.42	1.94	0.147 ± 3	0.1022 ± 32	1664 ± 59	–47
GGU 418739							
1.1*	1434	0.02	0.12	0.343 ± 7	0.1175 ± 10	1919 ± 15	–1
2.1	118	0.73	0.53	0.367 ± 10	0.1222 ± 22	1988 ± 33	1
3.1*	1494	0.17	0.23	0.361 ± 6	0.1212 ± 6	1973 ± 9	1

Table 2 *(continued)*

Spot	U (ppm)	Th/U	f_{206} (%)	$^{206}Pb/^{238}U$	$^{207}Pb/^{206}Pb$	Age	Disc. (%)
GGU 418777							
1.1*	1559	0.04	0.09	0.320 ± 5	0.1150 ± 6	1879 ± 10	–5
2.1	5098	0.10	0.02	0.372 ± 7	0.1237 ± 12	2011 ± 17	2
3.1*	1771	0.05	0.30	0.341 ± 6	0.1195 ± 5	1949 ± 7	–3
4.1*	752	0.15	0.28	0.328 ± 6	0.1145 ± 6	1871 ± 10	–2
5.1*	3079	0.16	0.03	0.342 ± 6	0.1204 ± 4	1962 ± 6	–3
6.1	92	0.51	1.11	0.354 ± 11	0.1209 ± 31	1970 ± 47	–1
GGU 418670							
1.1	294	0.55	0.17	0.353 ± 11	0.1219 ± 21	1984 ± 30	–2
2.1	177	0.35	0.34	0.356 ± 9	0.1232 ± 24	2003 ± 35	–2
3.1	232	0.36	0.18	0.342 ± 9	0.1215 ± 11	1978 ± 16	–4
4.1	720	0.30	0.51	0.115 ± 2	0.0955 ± 14	1538 ± 28	–54
5.1	66	0.42	0.18	0.338 ± 8	0.1253 ± 25	2033 ± 36	–8
6.1	80	0.46	0.36	0.347 ± 10	0.1204 ± 36	1962 ± 54	–2
7.1*	2030	0.03	0.25	0.352 ± 8	0.1186 ± 6	1936 ± 10	0
8.1*	1727	0.04	0.68	0.311 ± 6	0.1165 ± 10	1904 ± 15	–8
9.1	38	0.45	2.72	0.366 ± 19	0.1161 ± 67	1897 ± 107	6
GGU 418774							
1.1	180	0.19	1.11	0.224 ± 9	0.1105 ± 28	1807 ± 47	–28
2.1*	3115	0.03	0.46	0.081 ± 1	0.0705 ± 9	944 ± 25	–47
3.1*	1805	0.07	4.88	0.080 ± 2	0.0754 ± 22	1079 ± 59	–54
4.1	909	0.17	2.50	0.127 ± 3	0.0979 ± 20	1584 ± 39	–51
5.1*	1795	0.02	0.99	0.078 ± 1	0.0706 ± 16	945 ± 48	–49
6.1#	994	0.02	2.50	0.070 ± 1	0.0575 ± 25	509 ± 99	–15
7.1#	918	0.02	2.49	0.071 ± 1	0.0712 ± 24	963 ± 69	–54
8.1*	1226	0.01	0.80	0.069 ± 1	0.0626 ± 13	696 ± 43	–38
9.1	489	0.37	0.38	0.259 ± 6	0.1195 ± 16	1949 ± 23	–24
10.1	395	0.31	1.90	0.244 ± 6	0.1168 ± 22	1907 ± 34	–26
11.1	511	0.06	1.11	0.159 ± 4	0.1039 ± 16	1694 ± 29	–44
GGU 418705							
1.1	162	0.43	0.05	0.362 ± 7	0.1238 ± 11	2012 ± 16	–1
2.1#	157	0.39	0.40	0.362 ± 8	0.1211 ± 16	1973 ± 23	1
3.1	125	0.45	1.08	0.349 ± 7	0.1232 ± 24	2003 ± 35	–4
4.1	185	0.52	0.93	0.333 ± 10	0.1221 ± 21	1986 ± 31	–7
5.1	159	0.59	0.93	0.347 ± 8	0.1209 ± 21	1970 ± 31	–2

Spots marked with * represent rims; spots marked # are outgrowths; f_{206} is the proportion of ^{206}Pb that is not radiogenic; Age is the apparent $^{207}Pb/^{206}Pb$ age; Disc. is the degree of discordance between the $^{207}Pb/^{206}Pb$ and $^{206}Pb/^{238}U$ ages. All errors quoted at 1σ level.

The third sample (GGU 418719), a K-rich granitic biotite gneiss, yielded nearly exclusively metamict zircons and did not provide a reliable age. All three zircons analysed were discordant; one had a $^{207}Pb/^{206}Pb$ age of 2360 Ma and probably represents an inherited Archaean zircon which experienced severe Pb loss during one or more metamorphic events.

Granitoid sheets within basement gneisses

Four samples were studied of metaporphyritic granite sheets cutting basement gneisses. Two of these (GGU 419733, 432934; Fig. 1) yielded transparent zircons with ages around 1750–1800 Ma (three analyses each, Table 1).

Two other granite sheets (GGU 419802, 432808; Fig. 1) yielded only metamict zircons. Nearly opaque zircons were recovered from GGU 419802, a metagranite cutting paragneisses in the crystalline basement south of Lambert Land. Four grains were analysed, but no satisfactory age was obtained. Two analyses were nearly concordant: one, on a transparent part of a zircon, at *c*. 1970 Ma, and a metamict high-U grain (1350 ppm U) at *c*. 410 Ma ($^{206}Pb/^{238}U$ age); the remaining two analyses were very discordant, and one grain with a $^{207}Pb/^{206}Pb$ age of 2770 Ma is probably an inherited Archaean zircon.

Three analyses on zircons from GGU 432808 plot on a discordia line between 2000 and 450 Ma, but are too discordant to define a precise age.

Granitoid rocks associated with metaquartzitic sandstones in the Nørreland window

About 25 per cent of the area within the Nørreland window consists of felsic metaporphyritic rocks which appear to have intrusive relationships with the associated metasandstones, although strong deformation has made interpretation of the field observations difficult. Three samples were investigated.

A metaporphyritic quartz diorite (GGU 419967, Fig. 3B) yielded nearly opaque zircons with thin clear rims (too thin to be analysed). Three out of five zircons are close to concordant at *c*. 1860 Ma (Fig. 4B, Table 1); the two remaining grains yielded highly discordant data. All data points plot on a discordia line between 400 and 1890 Ma.

The meta-quartz diorite is cut by a metagranite sheet (GGU 419966) which yielded clear euhedral zircons with an upper intersection age of *c*. 1730 Ma (Fig. 4C); this we interpret as the age of emplacement of the granite.

A metagranodiorite (GGU 432871) from the Nørreland window yielded zircons with brown euhedrally zoned U-rich cores and wide clear rims. Five spot analyses (two cores and three rims) plot on a discordia line between 1900 and 400 Ma (Fig. 4D). The rims fall close to the upper intercept with concordia, while the cores plot nearer the lower intercept. The most plausible interpretation of this feature is that the upper intersection at *c*. 1900 Ma defines the emplacement of the granitoid rock, and that the scatter in isotopic compositions is due to variable Pb loss during Caledonian metamorphism, the metamict (more U-rich) cores (analyses 1.2 and 3.1, Table 2) having lost more Pb than the non-metamict rims.

Granitoid rocks from the Schnauder Ø complex

Two samples (GGU 418739 and 418777) were studied from the Schnauder Ø complex to test the possibility suggested by field observations that it could be of Caledonian age (Haller 1971; Jones & Escher 1995). Zircons from the two investigated samples are relatively clear and euhedral; a number of grains display distinct cores and rims. Three analyses were carried out on zircons from GGU 418739 including two rims, and six on zircons from GGU 418777, including four rims (Fig. 4E). All analyses are near-concordant and plot on or just below concordia between 1850 and 2000 Ma. Discordia calculations for these samples did not give meaningful results.

Granitoid rocks from Lambert Land and Hovgaard Ø

Three samples of granitoid rocks from Lambert Land and Hovgaard Ø were analysed, for which a Caledonian age was considered likely in the field either because they cut metadolerites lithologically similar to Midsommersø Dolerites or because of their occurrence within Caledonian shear zones.

Sample GGU 418670, a strongly deformed augen granite, was collected from a late Caledonian shear zone on southern Hovgaard Ø (Fig. 1), north of Lambert Land, to test whether it could have been emplaced during formation of the shear zone. Zircons vary from clear to metamict; many crystals are euhedral with sharp terminations, while some grains have distinct cores. Nine spots were analysed, two rims and seven cores. The seven analyses on cores plot close to a discordia with upper intersection at *c*. 1960 Ma. The two rims (one concordant) have high U concentrations and a very low Th/U ratio, suggesting formation during high-grade metamorphism (Williams & Claesson 1987); they have yielded $^{207}Pb/^{206}Pb$ ages of 1904 and 1936 Ma (Table 2).

A muscovite-biotite granite (GGU 418774, for composition see Fig. 3) was collected from a granite sheet within a presumed Caledonian thrust zone in the basement complex. This granite cuts across a Caledonian eclogite pod as well as its retrograde amphibolite border, suggesting that it might have been emplaced during Caledonian thrusting (Jones & Escher 1995). Most zircons from GGU 418774 are metamict to varying degrees and possess a complex structure: cores, more or

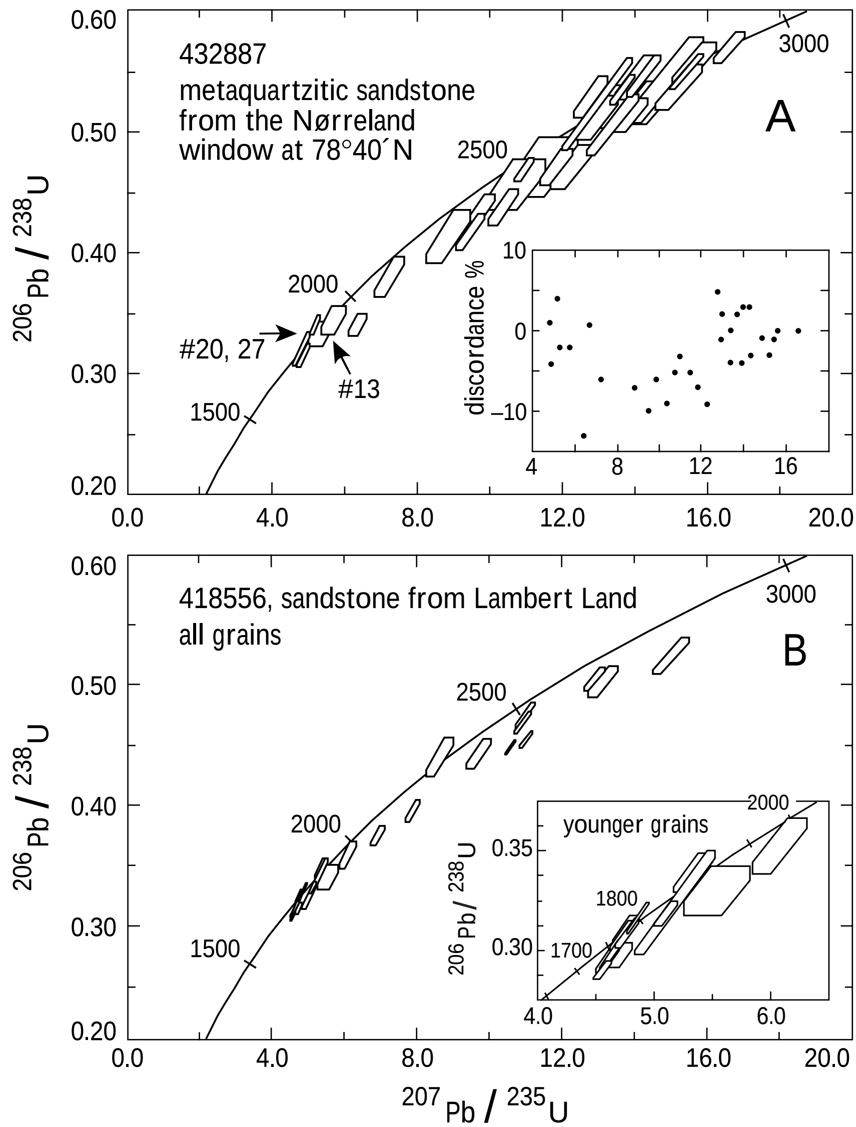


Fig. 5. Concordia diagrams for zircons from metasandstones, East Greenland Caledonian fold belt; error boxes display 1σ errors. For localities see Fig. 1.

less free of inclusions are surrounded by wide inclusion-rich rims which, in turn, have local inclusion-free protuberances ('outgrowths'). Eleven spots were analysed: four cores, four rims, one core overlapping rim, and two outgrowths. Most analyses yielded discordant data (Fig. 4F). They scatter along a discordia line with an upper intercept at *c.* 2000 Ma and lower intercept at *c.* 400 Ma. Rims and outgrowths plot near the lower intercept (Fig. 4G); cores fall more towards the upper intercept (Fig. 4F). There is a clear chemical difference between cores and rims (Table 2): cores having 180–510 ppm U and Th/U 0.06–0.37; rims 1230–3100 ppm U and Th/U 0.01–0.07. The outgrowths have intermediate values (900–1000 ppm U, Th/U 0.02). Two interpretations are consistent with the isotope data: either the upper intercept dates emplacement of the granite, and the scatter down to 400 Ma is due to strong Caledonian disturbance and new zircon growth, or GGU 418774 represents a Caledonian granite with inherited 2000 Ma zircons.

The last sample of this group, a deformed augen granodiorite (GGU 418705) cutting a mafic dyke, was collected on Lambert Land from an isolated outcrop in an area dominated by sandstones. The sample yielded clear to slightly turbid euhedral zircons with sharp terminations. No cores are present, but some grains have metamorphic (?) outgrowths. Five grains were analysed, including one outgrowth. All analyses are concordant or near-concordant and yield an age of *c.* 2000 Ma (Fig. 4H).

Table 3. SHRIMP U–Pb zircon ages for supracrustal rocks from the Caledonian fold belt

Spot	U (ppm)	Th/U	f_{206} (%)	$^{206}Pb/^{238}U$	$^{207}Pb/^{206}Pb$	Age	Disc. (%)
Metasandstones							
GGU 432887, Nørreland							
1.1	38	0.93	0.27	0.438 ± 15	0.1718 ± 31	2575 ± 31	–9
2.1	129	0.72	0.11	0.535 ± 12	0.1852 ± 13	2700 ± 12	2
3.1	79	0.80	0.38	0.529 ± 17	0.1753 ± 28	2608 ± 27	5
4.1	39	0.97	0.27	0.521 ± 18	0.1860 ± 47	2707 ± 42	0
5.1	46	0.70	0.02	0.471 ± 15	0.1821 ± 33	2672 ± 31	–7
6.1	15	1.02	0.02	0.471 ± 18	0.1889 ± 46	2733 ± 41	–9
7.1	101	1.31	0.06	0.570 ± 13	0.2111 ± 24	2914 ± 18	0
8.1	25	0.50	0.09	0.524 ± 18	0.1984 ± 67	2813 ± 56	–3
9.1	78	0.88	0.07	0.437 ± 11	0.1628 ± 33	2485 ± 34	–6
10.1	18	0.76	0.25	0.456 ± 21	0.1706 ± 87	2564 ± 88	–5
11.1	40	0.75	0.38	0.471 ± 25	0.1765 ± 106	2620 ± 104	–5
12.1	57	0.61	0.12	0.556 ± 18	0.2039 ± 37	2858 ± 30	0
13.1	47	0.52	0.47	0.333 ± 10	0.1158 ± 54	1892 ± 86	–2
13.2	28	0.75	0.34	0.344 ± 12	0.1199 ± 51	1955 ± 78	–2
14.1	130	0.68	0.02	0.550 ± 15	0.2036 ± 12	2855 ± 9	–1
15.1	62	0.55	1.45	0.340 ± 9	0.1356 ± 36	2171 ± 47	–13
15.2	59	0.42	0.02	0.380 ± 17	0.1382 ± 38	2205 ± 48	–6
16.1	72	0.93	0.02	0.546 ± 17	0.1891 ± 18	2734 ± 16	3
17.1	154	0.95	0.36	0.514 ± 14	0.1959 ± 33	2793 ± 28	–4
18.1	134	0.43	0.17	0.543 ± 36	0.1984 ± 42	2813 ± 35	–1
19.1	183	0.24	0.19	0.506 ± 25	0.1919 ± 24	2759 ± 21	–4
20.1	442	0.13	0.93	0.193 ± 4	0.1089 ± 20	1781 ± 34	–36
20.2	358	0.15	0.45	0.315 ± 10	0.1123 ± 15	1837 ± 24	–4
21.1	74	0.50	0.29	0.417 ± 15	0.1646 ± 24	2504 ± 25	–10
22.1	237	0.91	0.20	0.523 ± 38	0.1798 ± 16	2651 ± 15	2
23.1	49	0.69	0.35	0.537 ± 19	0.2055 ± 35	2870 ± 28	–3
24.1	104	0.81	0.17	0.675 ± 27	0.2563 ± 19	3224 ± 12	3
25.1	103	0.74	0.57	0.512 ± 18	0.1833 ± 46	2683 ± 42	–1
26.1	32	0.71	1.44	0.413 ± 22	0.1556 ± 55	2409 ± 62	–7
27.1	339	0.50	0.15	0.340 ± 8	0.1108 ± 12	1812 ± 19	4
27.2	325	0.48	0.08	0.320 ± 14	0.1089 ± 14	1780 ± 24	1
28.1	84	1.69	0.13	0.545 ± 18	0.1864 ± 26	2711 ± 23	3
29.1	121	0.30	0.07	0.469 ± 10	0.1695 ± 18	2552 ± 18	–3
GGU 418556, Lambert Land							
1.1	271	0.44	1.47	0.329 ± 21	0.1144 ± 13	1871 ± 20	–2
2.1	190	0.37	0.57	0.462 ± 9	0.1713 ± 10	2571 ± 10	–5
3.1	984	0.09	0.39	0.441 ± 6	0.1738 ± 5	2595 ± 5	–9
4.1	851	0.04	0.22	0.308 ± 5	0.1112 ± 13	1819 ± 22	–5
5.1	834	0.03	0.26	0.302 ± 4	0.1096 ± 7	1793 ± 12	–5
6.1	126	0.70	0.63	0.498 ± 10	0.1876 ± 16	2722 ± 14	–4
7.1	493	1.02	2.57	0.448 ± 7	0.1783 ± 9	2637 ± 9	–9
8.1	194	0.45	0.47	0.325 ± 5	0.1140 ± 10	1864 ± 16	–3
9.1	967	0.06	6.78	0.368 ± 8	0.1363 ± 18	2180 ± 23	–7
10.1	1510	0.04	0.14	0.318 ± 6	0.1083 ± 11	1771 ± 18	1
11.1	213	0.30	0.35	0.389 ± 9	0.1469 ± 15	2310 ± 18	–8
12.1	117	0.41	0.08	0.436 ± 12	0.1613 ± 25	2469 ± 27	–5
13.1	223	0.59	0.11	0.468 ± 11	0.1698 ± 13	2556 ± 13	–3
14.1	1117	0.04	2.01	0.306 ± 5	0.1089 ± 10	1782 ± 17	–3
15.1	175	0.95	0.15	0.517 ± 15	0.2105 ± 23	2910 ± 18	–8

Table 3 *(continued)*

Spot	U (ppm)	Th/U	f_{206} (%)	$^{206}Pb/^{238}U$	$^{207}Pb/^{206}Pb$	Age	Disc. (%)
16.1	107	0.19	0.50	0.341 ± 8	0.1136 ± 18	1858 ± 28	2
17.1	69	0.77	0.34	0.496 ± 13	0.1918 ± 27	2758 ± 23	–6
18.1	125	0.47	0.01	0.352 ± 11	0.1252 ± 24	2032 ± 34	–4
19.1	699	0.03	0.15	0.317 ± 6	0.1093 ± 7	1787 ± 12	–1
20.1	1138	0.05	0.22	0.308 ± 6	0.1091 ± 7	1785 ± 12	–3
21.1	1651	0.03	0.11	0.323 ± 6	0.1091 ± 3	1784 ± 6	1
22.1	46	1.34	0.95	0.334 ± 10	0.1204 ± 46	1963 ± 70	5
23.1	70	0.54	0.68	0.433 ± 16	0.1443 ± 28	2280 ± 34	2
24.1	1743	0.04	0.10	0.316 ± 6	0.1088 ± 10	1779 ± 17	–1
Metarhyolitic rocks							
GGU 423424							
1.1	742	0.61	0.05	0.278 ± 7	0.1052 ± 10	1718 ± 17	–8
2.1	38	0.77	1.22	0.304 ± 11	0.1008 ± 49	1638 ± 93	5
3.1	76	0.51	0.17	0.304 ± 10	0.1047 ± 30	1709 ± 53	0
4.1	793	0.84	0.01	0.284 ± 7	0.1055 ± 10	1723 ± 17	–7
5.1	52	0.99	0.08	0.300 ± 17	0.1094 ± 29	1789 ± 49	–5
6.1	46	0.75	0.35	0.304 ± 11	0.1062 ± 26	1735 ± 46	–1
7.1	199	1.00	0.02	0.307 ± 13	0.1072 ± 15	1753 ± 25	–2
7.2	170	0.97	0.03	0.309 ± 10	0.1077 ± 16	1760 ± 27	–1
8.1	70	1.10	0.00	0.306 ± 9	0.1098 ± 25	1796 ± 42	–4
9.1	511	0.73	0.03	0.298 ± 9	0.1065 ± 7	1741 ± 12	–3
10.1	1077	1.62	0.03	0.306 ± 8	0.1068 ± 4	1746 ± 7	–1
GGU 423425							
1.1	379	1.66	0.24	0.303 ± 8	0.1062 ± 10	1735 ± 17	–2
2.1	228	0.65	0.34	0.310 ± 6	0.1059 ± 18	1731 ± 31	0
3.1	341	0.71	0.21	0.308 ± 6	0.1065 ± 18	1741 ± 31	–1
4.1	311	0.81	0.27	0.313 ± 8	0.1062 ± 11	1734 ± 19	1
5.1	144	0.90	0.77	0.310 ± 6	0.1057 ± 23	1727 ± 41	1
6.1	448	1.13	0.04	0.307 ± 9	0.1065 ± 8	1740 ± 14	–1
7.1	402	0.18	0.11	0.307 ± 8	0.1056 ± 9	1725 ± 16	0
8.1	535	0.65	0.14	0.305 ± 5	0.1064 ± 7	1739 ± 12	–1
9.1	353	0.70	0.11	0.304 ± 6	0.1077 ± 11	1760 ± 20	–3
10.1	183	0.77	0.24	0.312 ± 8	0.1078 ± 25	1762 ± 43	–1

f_{206} is the proportion of ^{206}Pb that is not radiogenic; Disc. is the degree of discordance between the $^{207}Pb/^{206}Pb$ and $^{206}Pb/^{238}U$ ages; Age is the apparent $^{207}Pb/^{206}Pb$ age. All errors quoted at 1σ level.

Sandstones

Detrital zircons were investigated from two samples of metasandstone from the fold belt, traditionally correlated with the Independence Fjord Group: GGU 432887 from the Nørreland window, and GGU 418556 from Lambert Land.

The sample from Nørreland is a strongly strained quartzite. Nearly all zircons appear detrital; they are rounded and have pitted surfaces. Most grains have low U (< 100 ppm) and most analyses yielded Archaean $^{207}Pb/^{206}Pb$ ages with a significant scatter along a discordia line between *c.* 1800 and 2800 Ma (Fig. 5A). Analyses with a $^{207}Pb/^{206}Pb$ age between 1800 and 2700 Ma appear to be more discordant than those around 1800 and 2700–2900 Ma (Fig. 5A, inset), which suggests that most of these ages may have no chronological meaning but, more probably, are the result of Pb loss from Archaean zircons during a Palaeoproterozoic metamorphic event. One zircon (No. 13, Table 3,

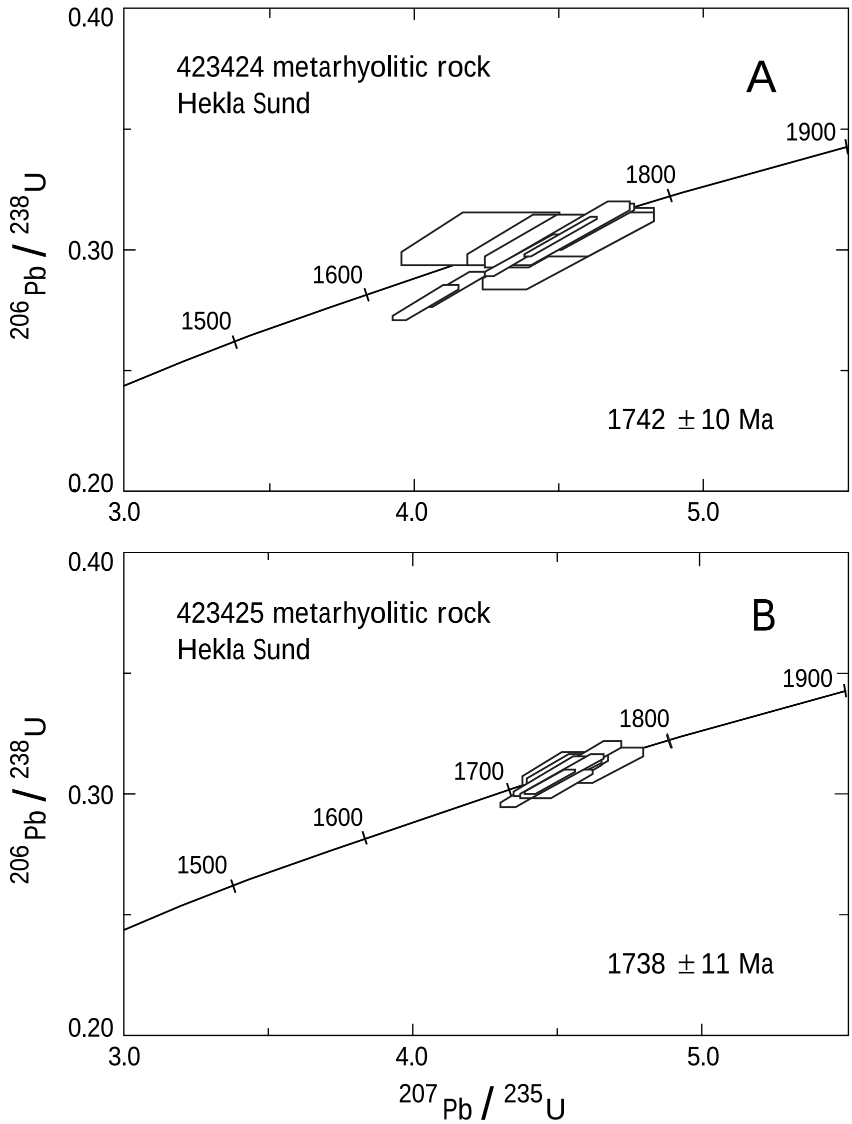


Fig. 6. Concordia diagrams for zircons from metarhyolitic rocks interlayered with metabasaltic pillow lavas at Hekla Sund; error boxes display 1σ errors. For locality see Fig. 1.

Fig. 5A), with a $^{207}Pb/^{206}Pb$ age of *c*. 1900 Ma, is concordant and probably represents a truly Proterozoic detrital zircon. Two grains (Nos 20 and 27, Fig. 5A) yielded concordant $^{207}Pb/^{206}Pb$ ages of about 1800 Ma. These grains have significantly higher U than all other analysed zircons and could be metamorphic in origin.

Most zircons from the Lambert Land sample (GGU 418556) are abraded and variably rounded and are thus clearly of detrital origin. $^{207}Pb/^{206}Pb$ ages for this group range from 1900 to 2900 Ma (Table 3, Fig. 5B). Most of these zircons yielded discordant data, and the ages cannot be taken at face value, but a few grains gave near-concordant ages between 1800 and 2050 Ma (Fig. 5B, inset) which we interpret as close to their true age. Another group of zircons is not obviously abraded; these grains are strongly coloured and show spectacular twinning in some cases. Eight grains were analysed; they have high U and low Th/U ratios (Table 3) and all yielded ages of about 1775 Ma (Fig. 5B, inset). If the strongly coloured zircons were derived from 1775 Ma granitic rocks their well-preserved structures suggest they were not transported far. The isotope data do not exclude the possibility that these grains represent metamorphic zircons formed within the sandstone, but in view of the twinning this would appear less likely to be the case.

Rhyolitic rocks

Two samples were studied of feldspar-phyric metarhyolitic rocks associated with the basalt sequence at Hekla Sund. As noted above, the basalt–rhyolite sequence is interlayered with sandstones correlated with the Independence Fjord Group, and forms part of the Caledonian thrust units. Zircons occur mainly as crystal fragments, up to 200 µm long, sometimes with euhedral terminations. Nine (out of 10) analyses on zircons from GGU 423424 and ten analyses on zircons from GGU 423425 yielded tight clusters on concordia at 1742 ± 10 and 1738 ± 11 Ma, respectively (Fig. 6, Tables 1, 2). Taken together the zircons from the two samples give a weighted mean age of 1740 ± 6 Ma (MSWD 0.52), which we interpret as dating the time of emplacement of the rhyolitic rocks and the associated basalts; this age has clear implications for the age of the sandstone sequence.

Discussion

Interpretation of the isotope data on the granitoid rocks

With one possible exception (GGU 418774) all investigated granitoid samples yielded Palaeoproterozoic zircon dates of 1730–2000 Ma, which normally would be interpreted as the age of igneous emplacement of the rocks. This does not agree with the impression gained during field work that some of the granitoid rocks could be of Caledonian age. However, granitoid rocks derived from crustal sources often contain zircons inherited from their parents (e.g. Pidgeon & Compston 1992), and the question has to be considered whether the analysed zircons could be inherited, and therefore give no information on the true age of the rocks. For several reasons we consider this as unlikely.

1. The granite for which a Caledonian age is most strongly supported by field observations (GGU 418774, which cuts through a Caledonian eclogite pod as well as its retrogressed rim), is also the only sample where zircon rims plot close to 400 Ma on a discordia line (Fig. 4G). Zircons from most other samples, cores as well as rims, plot on or close to concordia at 1750–2000 Ma (e.g. GGU 418777 and 418705; Fig. 4E, H). If these samples represent Caledonian granites, *all* analysed zircons would have to be inherited, whereas zircons formed during Caledonian granite formation are so rare that they were not detected. This is unlikely: Caledonian granites from Scotland, for example those studied by Pidgeon & Compston (1992), have both magmatic and inherited zircons present in significant proportions.

2. Inherited zircons in granitic rocks commonly yield widely variable ages (e.g. Pidgeon & Compston 1992). In the samples investigated here all zircons have more or less the same age (e.g. GGU 418705, Fig. 4H), which is less likely to be the case if they represent anatectic granites containing only inherited zircons.

3. Palaeoproterozoic zircons in several of the investigated samples are perfectly euhedral, with well preserved sharp pyramidal terminations (e.g. GGU 418705). Such features are unlikely to have survived Caledonian melting, since at least part of the zircon would probably go into solution (Watson & Harrison 1983).

4. Inherited zircons are common in anatectic granites, especially those formed at relatively low temperatures; they occur most frequently in S-type granites which are typically formed at temperatures of 700–750°C, compared to 800–900°C for I-type granitoids (Watson & Harrison 1983). The two samples within which inherited (Archaean) zircons have been found (GGU 418719 and 419802, see above) belong indeed to the group of low-temperature leucogranites shown in Fig. 3, but several of the other samples studied are more mafic I-type granodiorites or tonalitic rocks (for example GGU 418705 and 432871, Fig. 3), within which inherited zircons are much less common.

Together, we consider these features are very strong evidence that (with the possible exception of GGU 418774) the investigated granites were emplaced during Palaeoproterozoic, not Caledonian, orogenic events.

Interpretation of the isotope data on the sandstones

In view of the preponderance of Palaeoproterozoic gneisses and granitoid rocks in the region it is surprising that detrital 1750–2000 Ma zircons are relatively rare in the investigated sandstones from the Caledonian fold belt. Most zircons analysed are Archaean, but appear to have experienced severe Pb loss during high-grade Palaeoproterozoic metamorphism. Large areas in West Greenland and southern East Greenland are composed of Archaean rocks reworked during Palaeoproterozoic tectonothermal events around 1800–1850 Ma (see the Geological map of Greenland, 1:2 500 000, Escher &

Pulvertaft 1995). Such rocks also occur (at least locally) beneath the Inland Ice (Escher & Pulvertaft 1995; Weis *et al*. 1997). It is plausible that most zircons in the investigated metasandstones were derived by erosion of such a terrane.

Deposition of the sandstones probably took place after the end of the Palaeoproterozoic igneous and tectono-metamorphic events, in agreement with the undeformed and unmetamorphosed nature of the Independence Fjord Group sandstones in the Caledonian foreland. The presence of a few detrital zircons of Palaeoproterozoic age (1800–2050 Ma) in the sandstones of the Caledonian fold belt supports this suggestion. Since the 1740 Ma basalt sequence at Hekla Sund is interlayered with the metasandstones of Lambert Land and Kronprins Christian Land (Fig. 1), the age of the basalts provides a time point within the period of deposition of the sandstones. Deposition thus must have started already during the later Palaeoproterozoic, very soon after the end of the Palaeoproterozoic orogenic events in the region. Since any metamorphism of the sandstones must have taken place later than *c*. 1740 Ma, the strongly coloured, U-rich, 1775 Ma zircons in the sample from Lambert Land are probably detrital rather than *in situ* metamorphic in origin, which would constrain deposition of at least some of the sandstones on Lambert Land to the period 1775–1740 Ma. This is much earlier than anticipated: on the Geological map of Greenland (Escher & Pulvertaft 1995) the sandstones of Lambert Land and their assumed correlatives of the Independence Fjord Group in the Caledonian foreland are shown as Mesoproterozoic deposits.

Deposition of some of the sandstones around 1750 Ma ago would permit the possibility that they were intruded by the youngest of the dated granite sheets in the region. This may be the case in the Nørreland window where the quartzitic sandstones appear to be cut by granites. One of these (GGU 419966) has yielded an age of about 1730 Ma (Table 1), and may have been emplaced into very young sandstone. However, most of the granites (e.g. GGU 418705 in Lambert Land) are much older than the sandstones with which they are now spatially associated. Lambert Land is characterised by a complex history of Caledonian thrusting (Jones & Escher 1995); this thrusting has apparently brought rocks of different ages into close juxtaposition.

Why are there no Caledonian granites in the northern part of the Caledonian fold belt?

Formation of granitic magmas in collisional orogens is dependent on several factors, the most important being: (1) the temperatures reached during orogenic thickening and later extension, and (2) the presence of 'fertile' crustal lithologies (e.g. Vielzeuf *et al*. 1990; Brown 1994). The metamorphic temperatures estimated for the eclogites in the Caledonian fold belt are 700–800°C (Brueckner *et al*. 1998; K.A. Jones, unpublished data), which are less than required for the formation of I-type granitoid magmas (Watson & Harrison 1983, p. 303). In the area under consideration the Precambrian basement consists mainly of 'sterile' Palaeoproterozoic granitoid rocks, with relatively small amounts of biotite and hornblende (most commonly < 10%); these do not permit large proportions of granitic magma to be formed by dehydration melting at temperatures below *c*. 900°C, because only small amounts of water will be generated by dehydration (e.g. Clemens & Vielzeuf 1987). 'Fertile' lithologies, such as pelitic metasedimentary rocks are rare within the crystalline basement. Those metasedimentary units that are present comprise the quartz-rich sandstones of Lambert Land and Nørreland, which would not easily melt, and large units of Neoproterozoic to lower Palaeozoic semipelitic lithologies that are present in high-level Caledonian thrust units in Kronprins Christian Land (Fig. 1). Those, however, are at a very low metamorphic grade, and have never been very deeply buried.

In contrast to the northern part of the Caledonian fold belt investigated during the present study, Caledonian granites are common in the southern part of the Caledonian fold belt, south of Bessel Fjord (76°N). The Geological map of Greenland, 1:2 500 000 (Escher & Pulvertaft 1995) shows that large parts of that region are underlain by high-grade metasedimentary sequences. Furthermore, many of the Caledonian granites occur within units of metasedimentary rocks, and formation by anatexis of the metasediments is suggested by field observations and geochemical data (Kalsbeek *et al*. 1998). The rarity of corresponding metasedimentary units in the northernmost part of the Caledonian fold belt may be the main reason why Caledonian granites do not occur in this area.

Acknowledgements

We thank Professor W. Compston, Australian National University, for permission to use SHRIMP 1 for this study, Peter Venslev (Copenhagen) and Neil Gabbitas (Canberra) for assistance with the sample preparation, and Niels Henriksen, Stuart Watt and A.K. Higgins for critical reading of earlier versions of the manuscript.

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